# Environmental changes close to the Ordovician–Silurian boundary

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## Synopsis

Most late Ordovician to early Silurian sequences show evidence of a regressive phase followed by transgression, reflecting glacio-eustatic sea-level changes. Continental glacial deposits are particularly well known from Saharan Africa, and glaciomarine deposits from Iberia and Normandy. Rapid growth of the ice caps at the beginning of the Hirnantian is reflected on clastic marine shelves by a change from mudstones to a variety of shallow marine sand facies. Withdrawal of the sea to the edges of shelves fed sand into basins to form submarine fans. Shallow carbonate shelves generally became exposed during the Hirnantian, and karstic surfaces developed. A sea-level fall of between 50 and 100 m is envisaged. The regressive deposits are usually abruptly overlain by deeper-water deposits formed during a rapid transgression. Graptolitic shales are widely developed on clastic shelves, but there is a return to shallow marine limestones on carbonate shelves. There is local evidence of oscillations of sea-level within the main Hirnantian glacial event, but it is uncertain whether these changes were eustatically controlled. It is suggested that the climate during the Hirnantian remained cold in peri-polar regions, but may have been variable in mid-latitudes and was tropical in equatorial regions. There is some palaeomagnetic evidence to suggest that continents were moving unusually fast during late Ordovician times, which might have had an influence on the growth and decay of late Ordovician ice caps.

## Introduction

Most late Ordovician to early Silurian sequences show evidence of a regressive phase followed by transgression. The regressive–transgressive interval is of the same age on plates which were separate in the Lower Palaeozoic (Berry & Boucot 1973) and so satisfies the criteria for identifying eustatic sea-level changes (Fortey 1984). The fall in sea-level started at the beginning of the Hirnantian and the subsequent rise of sea-level had been largely completed before the end of Hirnantian times. A major ice cap was present on the Gondwana plate at this time and it is likely that the sea-level changes were related to the growth and decay of that ice cap.

The Ordovician–Silurian boundary, as it is now placed at the base of the *P. acuminatus* Zone, post-dates the late-Ordovician sea-level changes and falls within a period of environmental stability. Thus the often striking facies changes in the Hirnantian, and particularly the change from shallow to deeper water facies at the top of the Hirnantian, help to identify horizons immediately below the boundary between the systems, but not the boundary itself.

## Duration of the eustatic changes

Different ways of estimating the duration of Hirnantian environmental changes can be made, and these produce somewhat different results. Estimates of the duration of the Hirnantian made by dividing the duration of the Ashgill, based on radiometric age determinations, by the number of stages (four) give 1·8 to 2·5 my. If the duration of the Ashgill is divided by the number of zones in the type area (eight) (Ingham 1966) the duration of the Hirnantian, which has only one zone, is 1 to 1·25 my. A value between 1 and 2 million years is probable, but more radiometric dates close to the Ordovician–Silurian boundary are needed to give more accurate estimates.

## Changes in sedimentary environments

**Continental glaciation.** The deposits of continental ice sheets of upper Ordovician age in Saharan Africa are well known through the descriptions of Beuf *et al.* 1971, Rognon *et al.* 1972, and others. They recognized nearly all the features characteristic of land-based ice deposition, including glaciated pavements, striated pebbles, tillites, varved sediments and dropstones, and a wide variety of fluvio-glacial sediments (Fig. 2, section 1), some of which are associated with long esker-like ridges. Similar deposits have been recognized in South Africa (Rust 1982), and glacial deposits believed to be of a similar age have been described from west Africa, South America (see Spjeldnaes 1981 and references therein) and Saudi Arabia (McClure 1978). The late Ordovician Gondwana glaciation was clearly of continental dimensions and appears to have extended from the south pole through at least 40° of latitude. There is no evidence of a contemporary ice cap in the Ordovician northern hemisphere, which, according to palaeogeographic reconstructions, had no continental areas near the pole at that time.

**Glaciomarine environments.** Tilloids of glaciomarine origin were initially identified by Dangeard & Doré (1971) in Normandy, and by Hempel & Weise (1967) in Thuringia. Subsequently, glaciomarine sediments, usually consisting of pebbly mudstones, have been recognized in Brittany (Hamoumi *et al.* 1980), Celtiberia (Carls 1975), west central Spain (Robardet 1981) and Portugal (Romano & Diggens 1973–74; Young 1985).

Most of the clasts in the tilloids can be matched with carbonate or coarse clastic horizons in the underlying succession, indicating that at times the ice was grounded and caused erosion. Striated clasts are recorded from Normandy (Dangeard & Doré 1971) and Navatrasierra, western Spain (personal observation). Deposition, however, appears to have been from floating ice, as indicated by the delicately laminated nature of some of the sediments, the presence of dropstones in Brittany (Hamoumi 1981), but above all by the nature of the predominantly massive sandy mudstones which lack associated sand deposits of fluvioglacial origin. In Spain and Portugal there is evidence of regression and emergence prior to the deposition of the tilloids (Fig. 2, sections 2 and 3), and there are variable proportions of normal marine sediments interbedded with the glaciomarine sediments.

At the time of the maximum continental glaciation of the Gondwana plate, the adjacent Armorican plate apparently lacked a continental ice sheet. Here, ice was locally grounded on recently exposed shelf sediments but at times of slightly higher sea-level there was widespread floating ice from which was deposited the mainly structureless sandy mud with its dispersed clasts.

**Clastic shelves.** Many sequences which formed on clastic shelves show an upward passage from mudstones to shallow marine sandstones. On shelves where there was an adequate supply of sand complete upward-coarsening regressive sequences were formed starting with shelf muds and passing gradationally upwards through various shoreface facies (Fig. 1, section 1), or sometimes more abruptly into a variety of shallow marine facies (Fig. 1, sections 2 and 3). At other places where there was channelling of the shelf, massive or cross-stratified sandstones lie with a sharp erosional base on the underlying sediments (Fig. 1, section 4, might represent such a situation). When a clastic shelf or relatively shallow basin was relatively starved of sediment the regressive sequence is condensed, sometimes to as little as a metre, and may be partly calcareous, as in Västergötland (Fig. 1, section 5) where there is a thin oolite bed, or in the Yangtze Basin where a thin bioclastic limestone caps graptolite shales (Fig. 1, section 6).

At most places shallow marine sediments of the regressive phase are succeeded abruptly by mudstones with a benthic fauna indicating a deep shelf environment, or by graptolitic shales. Facies formed during the rise of sea-level are usually less than a metre thick, suggesting that the transgression was rapid.

**Clastic basins.** There is evidence from the Welsh Basin that the end Ordovician regression caused sediments to be carried across the marginal shelves and produced an influx of coarse clastics into previously mainly argillaceous basin environments. Pebbly mudstones of mass flow origin, thick-to-thin bedded turbidites, some of which are channelled, and some slumped units suggest the presence of substantial base-of-slope fans (Fig. 1, sections 7 and 8). At the north-west margin of the basin, fan sediments with resedimented ooids and fragmented valves of a *Hirnantia* fauna overlie trilobite-bearing mudstones, suggesting that this particular fan accumulated at no great depth.

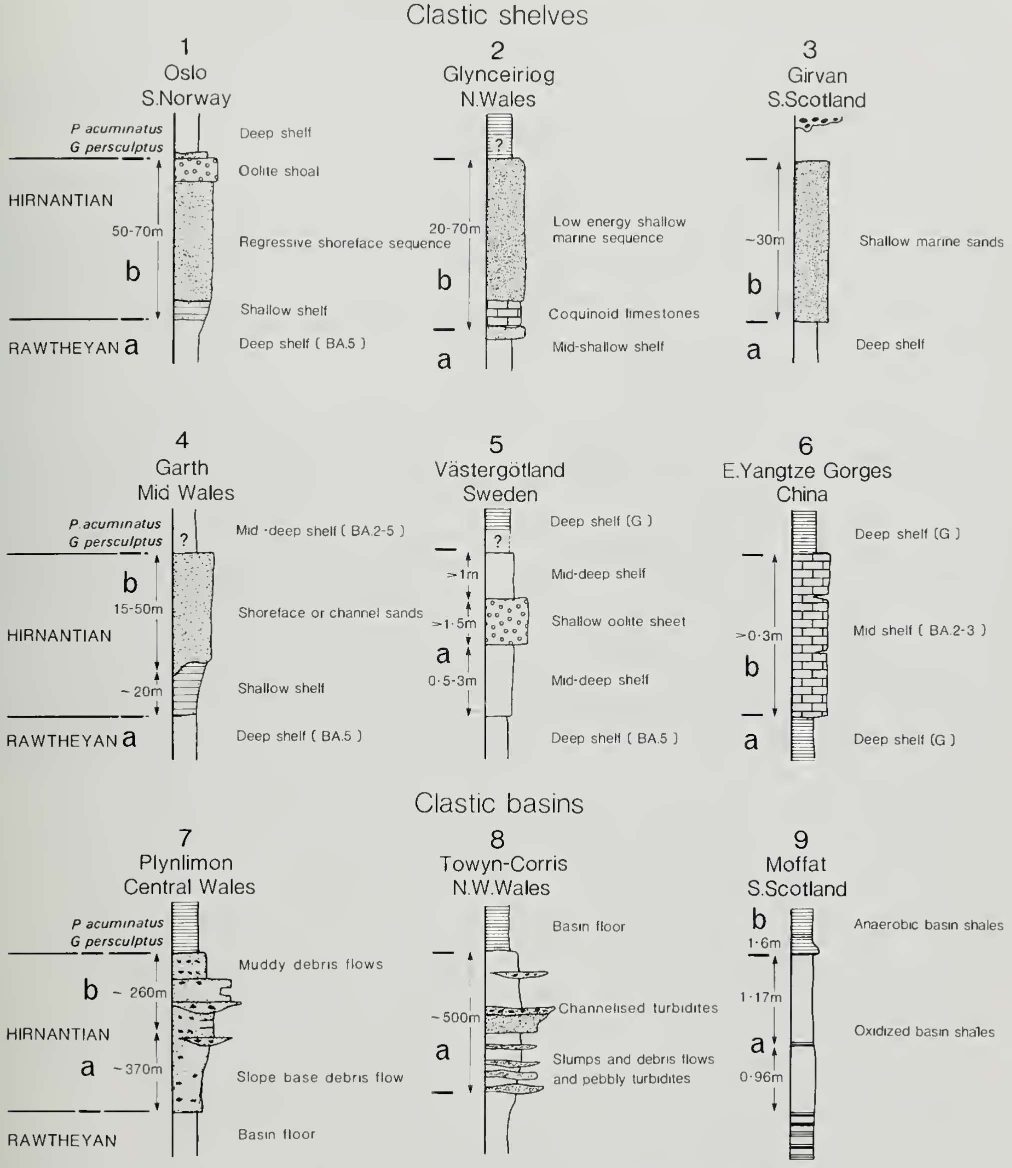


**Fig. 1** Generalized sections to show the sequence of environmental changes near the Ordovician/Silurian boundary. Data for the interpretations are to be found in the following references. Section 1: (a) Husbergøya Shale, (b) Langøyene Sandstone; Brenchley & Newall 1980. 2: (a) Dolhir Formation, (b) Glyn Formation; Hiller 1981; Brenchley & Cullen 1984. 3: (a) Drummuck Group, (b) High Mains Formation; Harper 1981. 4: (a) Wenallt Formation, (b) Cwm Clŷd Formation; Williams & Wright 1981. 5: (a) Dalmanitina Beds; Stridsberg 1980. 6: (a) Wufeng Formation, (b) Guanyinqiao Formation; Geng Liang-yu 1982. 7: (a) Nant-y-Moch Formation, (b) Drosgol Formation; James 1971; Cave 1979; James 1983. 8: Garnedd-Wen Formation; James 1972; James 1985. 9: (a) Upper Hartfell Shale Formation, (b) Birkhill Shale Formation; Williams 1983.

In some basins which were isolated from a source of coarse clastics there were no obvious changes in pelagic sedimentation, as in some of the graptolitic shale sequences in the Yukon (Lenz 1982; Lenz & McCracken 1982). In a rather similar graptolitic shale sequence at Dob's Linn in the Southern Uplands of Scotland, the end Ordovician regression cannot be identified but the transgression is reflected in a change from grey mudstones, without graptolites, to black graptolitic shales (Fig. 1, section 9). This change from oxidized to anoxic sediments might reflect the change from the vigorous bottom circulation of the glacial period to the more sluggish circulation following the melting of the ice caps.

The graptolitic shales, which commonly succeed the coarser clastics formed during the regression in basin environments, may contain a *G. persculptus* fauna, but may in other instances have *P. acuminatus* or even younger faunas in the lowest horizons. The local absence of the lowest Silurian graptolite zones is probably the result of erosion or non-deposition. Similar hiatuses are being increasingly recognized in DSDP cores in areas of pelagic sedimentation (Moore *et al.* 1978). For example, widespread deep-sea erosion in the Miocene is associated with periodic cold-climate events, lower eustatic sea-level and an intensification of bottom circulation (Keller & Barron 1983).

**Carbonate shelves.** Most of the very extensive carbonate platforms in North America and Arctic Canada appear to have been exposed at the end of the Ordovician, producing regional disconformities (Lenz 1976, 1982). The sedimentological effects of the regressive–transgressive cycle are commonly not easily recognized in shallow marine carbonate sequences. Nevertheless a late Ordovician, generally regressive, sequence culminating in a widespread oncoid bed has been recognized in Anticosti Island (Petryk 1981*a*), and this is succeeded by generally transgressive sediments (Fig. 2, section 5). At Manitoulin Island, Ontario, two karstic horizons separated by 15 cm of sediment occur close to the Ordovician–Silurian boundary in a sequence of shallow marine carbonate facies (Fig. 2, section 4). The effects of the end-Ordovician regression can also be recognized in the more offshore facies associated with carbonate mud mounds. In two of the carbonate mounds of the Boda Limestone (central Sweden) there is evidence of emergence of the mound crests, with karst surface on one mound (Fig. 2, section 6), and dripstone calcite lining fissures in the other. Graptolitic shales, formed after the transgression, mantle the


**Key** to Figs 1–2.

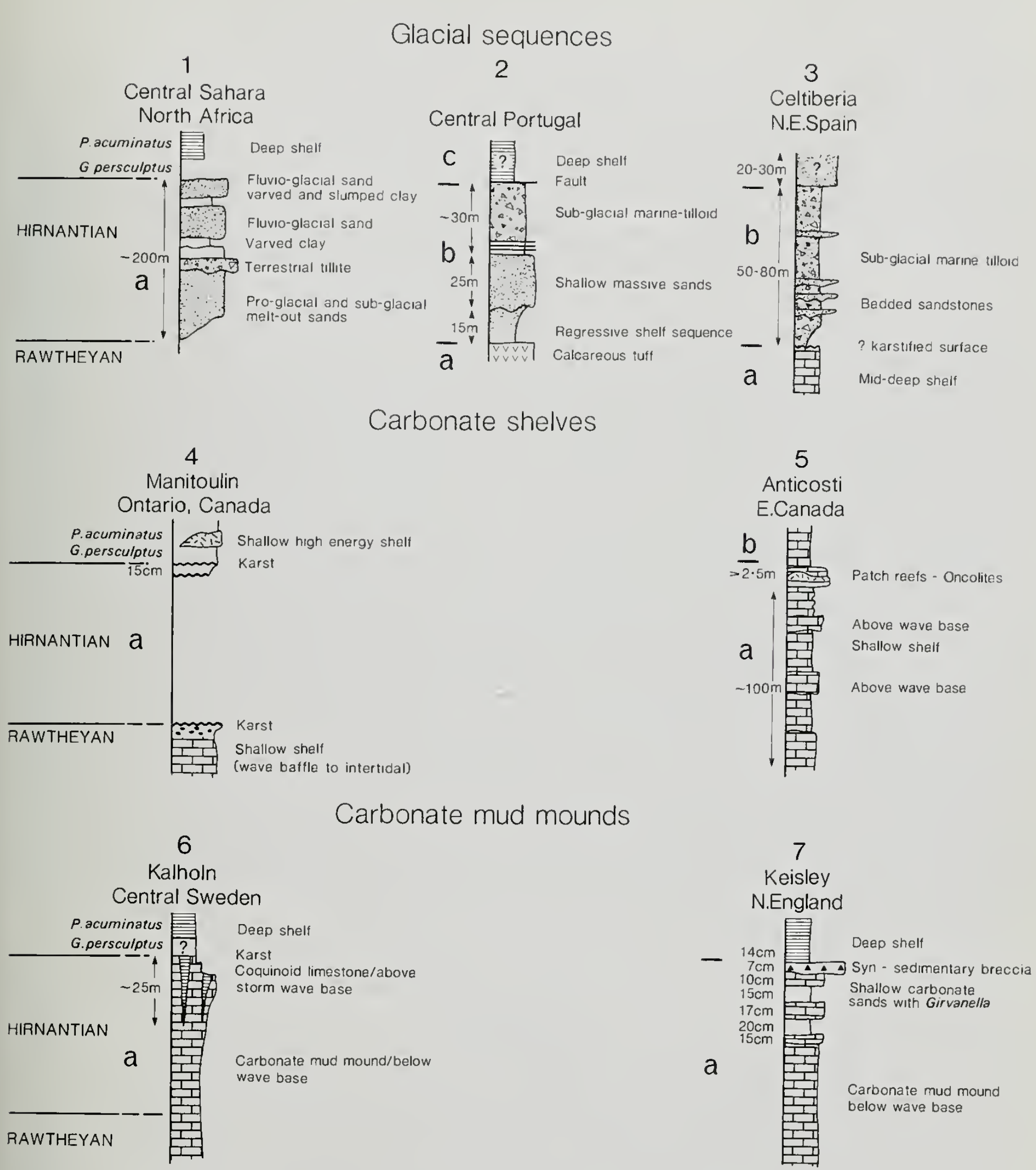


**Fig. 2** Generalized sections to show the sequence of environmental changes near the Ordovician–Silurian boundary. Data for the interpretations are to be found in the following references. Section 1: 'Unit IV'; Beuf *et al.* 1971. 2: (a) Porto de Santa Anna Formation, (b) Ribeira do Bracal Formation, (c) Ribeira Cimeria Formation; Young 1985. 3: (a) Cystoid Limestone, (b) Orea Shale; Carls 1975. 4: Georgian Bay Formation, (b) Manitoulin Formation; Copper 1978; Kobluk 1984. 5: (a) Ellis Bay Formation (up to Oncolites), (b) Becscie Formation; Petryk 1981*a*, 1981*b*. 6: Boda Limestone; Jaanusson 1979; Brenchley & Newall 1980. 7: Keisley Limestone; Wright 1985.

mounds and fill fissures in both cases. In the carbonate mound at Keisley, in northern England, the regression is reflected by the development of beds containing the alga *Girvanella* at the top of the mound. There is a final capping of breccia, a few cm thick, and this is succeeded abruptly by graptolitic shales, again marking the transgressive phase (Fig. 2, section 7).

**Bathymetric changes.** There is good evidence that most carbonate and clastic platforms and shelves shoaled to near sea-level or became exposed during the Hirnantian regression. Some of the platforms were already shallow before the start of the regression, but some muddy shelves which were initially below storm wave-base, suggesting water depths of several tens of metres, also became exposed (Brenchley & Newall 1980). The relief on an erosion surface below the Silurian in Iowa, USA, suggests that sea-level dropped at least 45 m (Johnson 1975). The emergence of the crests of carbonate mud mounds and the lining of fissures to a depth of nearly 30 m implies a sea-level fall of about 70 m (Brenchley & Newall 1980). A sea-level fall between 50 and 100 m seems likely though a figure of 'not more than 20 m' has been suggested by Geng Liang-yu (1982).

The widespread presence of grey mudstones with deep shelf benthic faunas prior to the regression, but graptolitic shales after the transgression, suggests that the sea-level rise might have been greater than its fall (Brenchley & Newall 1980). However, the evidence from carbonate platforms does not support this because in general early Silurian carbonates are similar to those of the late Ordovician and both suggest shallow marine environments. It may be that the development of early Silurian graptolitic facies is determined more by the preceding transgression which drowned many source areas, rather than by a substantial increase in water depths.

Although only a single regressive phase followed by transgression is apparent in many sections there is some evidence for oscillations of sea-level within the Hirnantian. Two karstic horizons representing two phases of emergence were recognized at Manitoulin Island (Kobluk 1984) and in a carbonate sequence near Oslo (Hanken 1974). Three regressive phases were identified by Petryk (1981*b*) in the upper Ordovician sequence on Anticosti Island. It is possible that these bathymetric changes might be related to phases of growth of the continental ice caps reflected by three separate horizons of till in the Saharan and South African sequences. Episodes of ice advance and retreat are now well documented in the Pleistocene record. Changes in the size of the Pleistocene ice caps produced cyclic changes in the $^{18}O/^{16}O$ isotopic record in oceanic sediments implying temperature fluctuations with a periodicity of about 20 000, 40 000 and 100 000 years (Hays *et al.* 1976) similar to those predicted by Milankovitch (1938) on astronomical grounds. A similar cyclicity might be expected in earlier glaciations, and might be represented by the three sea-level oscillations and three tills in the Hirnantian. However, the time-scale of these oscillations is still unclear.

**Geochemical changes.** There are very few studies of sediments close to the Ordovician–Silurian boundary which might show if the geochemistry reflected the climatic and other environmental changes. A pilot study in a relatively uniform sequence of argillaceous sediments in the type Ashgill area of northern England showed changes in carbonate, Fe and P content and in $Fe_2O_3$ activity at the base and/or top of the Hirnantian, which were correlated with minor changes in lithology and probably with changes in palaeobathymetry (Brenchley 1984). A study of carbon and oxygen stable isotopes in a sequence through a Boda carbonate mud mound showed changes in $^{18}O$ values which suggested a fall in sea-water temperature during the Hirnantian (Jux & Manze 1979). Both these studies suggest that further geochemical work might prove valuable in determining changes in sea-water chemistry and temperature during the Hirnantian.

**Climatic changes.** The distribution of late Ordovician glacial deposits suggests that continental ice sheets extended from the south pole through at least 40° of latitude and that there was floating ice for another 10° of latitude. The temperature of peripolar oceans would have been substantially depressed during such periods of glaciation. The effect of glaciation on the temperature of surface waters in lower latitudes is less easy to predict. Studies of surface waters at 18 000 years B.P., during the last interglacial, show marked differences between the Atlantic and Pacific Oceans, indicating there is no simple global pattern of temperature (McIntyre *et al.*

1976; Moore *et al.* 1980). Two points possibly relevant to the reconstruction of Ordovician climate do however emerge; one is that water temperatures in some tropical and temperate areas may actually be raised during a glacial episode, and the second is that notably cooler waters can develop in both temperate and tropical areas.

The widespread extension of cooler surface waters during a glaciation might explain the very broad distribution of the *Hirnantia* fauna, thought by some to be a cool-water fauna, throughout most temperate and sub-tropical regions during some part of Hirnantian times.

The possibility of elevated temperatures during a glacial phase might partly account for the apparently anomalous occurrence of Hirnantian oolitic horizons in sequences which were hitherto wholly clastic (Oslo in Norway, and Garth and Bala in north Wales). It is not necessarily a contradiction that the sequences which contain oolites also contain an *Hirnantia* fauna, since the changes of sea-surface temperatures can be substantial between glacials and interglacials, particularly in mid-latitudes.

A tentative construction of Hirnantian climate is that polar and peri-polar regions remained cool to glacial throughout the Hirnantian, mid-latitudes had very variable climatic conditions varying in time and space from cool to warm, while tropical areas in general remained hot. The climate instability and geographic contrasts of the Hirnantian were succeeded by more stable conditions in the Silurian. It is thought that the climate was in general similar to that of today, but that climatic belts were more nearly parallel to lines of latitude because of the relative absence of land in low latitudes (Ziegler *et al.* 1977).

**Palaeomagnetism.** The distribution of continents, based on palaeomagnetic evidence, has been reconstructed for the middle Ordovician and for the early Silurian (Ziegler *et al.* 1977; Ziegler & Scotese 1979; Scotese *et al.* 1979). Unfortunately there are no maps of comparable detail for the Upper Ordovician. Early Silurian reconstructions show Gondwanaland lying in high southern latitudes and other continents spread across the southern hemisphere and into mid-northern latitudes. No continents are located in high northern latitudes.

There is some evidence from the shape of the apparent polar-wandering paths of the Ordovician that the continents must have moved unusually fast in late Ordovician times, to create the Lower Silurian palaeogeography. Some confirmation of this rapid movement comes from a wealth of palaeomagnetic data in north and west Europe, which shows upper Ordovician (Caradoc and Ashgill) magnetism with steep inclination, implying a new polar position, contrasting with earlier and later data with significantly lower inclinations (Piper, 1987). Palaeomagnetic data from China also shows uppermost Ordovician poles differed in position from those earlier and later (Wang Xiaofeng *et al.* 1983). If these proposed unusually high rates of continental movement are confirmed they could have a significant bearing on the growth and decay of the late Ordovician ice caps (Piper, 1987).

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